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Coarse sediment transport in mountain streams in Colorado and Wyoming, USA

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Abstract

Since the early 1990s, US Forest Service researchers have made thousands of bedload measurements in steep, coarse-grained channels in Colorado and Wyoming, USA. In this paper we use data from 19 of those sites to characterize patterns and rates of coarse sediment transport for a range of channel types and sizes, including step–pool, plane-bed, pool–riffle, and near-braided channels. This effort builds upon previous work where we applied a piecewise regression model to (1) relate flow to rates of bedload transport and (2) define phases of transport in coarse-grained channels. Earlier, the model was tested using bedload data from eight sites on the Fraser Experimental Forest near Fraser, Colorado. The analysis showed good application to those data and to data from four supplementary channels to which the procedure was applied. The earlier results were, however, derived from data collected at sites that, for the most part, have quite similar geology and runoff regimes. In this paper we evaluate further the application of piecewise regression to data from channels with a wider range of geomorphic conditions. The results corroborate with those from the earlier work in that there is a relatively narrow range of discharges at which a substantial change in the nature of bedload transport occurs. The transition from primarily low rates of sand transport (phase I) to higher rates of sand and coarse gravel transport (phase II) occurs, on average, at about 80 per cent of the bankfull (1·5-year return interval) discharge. A comparison of grain sizes moved during the two phases showed that coarse gravel is rarely trapped in the samplers during phase I transport. Moreover, the movement and capture of the D_{16} to D_{25} grain size of the bed surface seems to correspond with the onset of phase II transport, particularly in systems with largely static channel surfaces. However, while there were many similarities in observed patterns of bedload transport at the 19 studied sites, each had its own 'bedload signal' in that the rate and size of materials transported largely reflected the nature of flow and sediment particular to that system. Published in 2005 by John Wiley & Sons, Ltd.

Keywords: **bedload transport; gravel-bed channels; piecewise regression**

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Introduction

A number of types of gravel-bed channel exist, ranging from steep, mountain channels with step–pool topography, to flatter channels in wide valley bottoms with pool–riffle or braided forms (Rosgen, 1994; Montgomery and Buffington, 1997). Nearly all gravel-bed channels have in common an armour layer composed of relatively coarse grains (large gravel, cobbles, and boulders) that overlie a finer subsurface, consisting primarily of sand and gravel. Different types of armour layers have been described and a number of mechanisms of formation have been suggested. For instance, a static armour layer is stable under most flows and may represent supply-restricted conditions (Sutherland, 1987) whereas a mobile armour layer is moved, though its structure maintained, under moderately high flows (Powell, 1998). The armour layer can be a relatively dynamic feature, responding to large-scale changes in flow regime or sediment supply (Reid *et al*., 1985; Dietrich *et al*., 1989; Buffington and Montgomery, 1999).

The presence of an armour layer is thought to suppress rates of bedload transport until a critical threshold is achieved. At least two phases of bedload transport, with notably differing qualities, have been described for gravel-bed streams (e.g. Jackson and Beschta, 1982; Carling, 1988; Ashworth and Ferguson, 1989; Warburton, 1992; Ryan *et al*., 2002). Phase I consists primarily of sand and small gravel moving at relatively low rates over a stable channel surface. The source of sediment in transport at lower flows is primarily from patches of finer grained materials (Lisle, 1995; Laronne *et al*., 2001) or pools (Lisle and Hilton, 1992, 1999; Trush *et al*., 2000). During phase II, more coarse grains are moved and the transport becomes more vigorous and irregular. Transition from phase I to phase II indicates disturbance and initiation of transport of grains constituting some portion of the armour layer and includes material from both the bed surface and subsurface. While the existence of different phases of transport is generally acknowledged, the threshold between them is often poorly defined, though flows near bankfull discharge are generally a first approximation (Buffington, 1995). Speculatively, the relative discharge at which the onset of phase II transport occurs may vary between channels with different types of armour layers.

Ryan *et al*. (2002) evaluated the application of a piecewise regression model for defining phases of bedload transport. Briefly, the analysis recognized the existence of different transport relationships for different ranges of flow. In that report, two linear segments, one for phase I and one for phase II, were defined for eight coarse-grained channels on the Fraser Experimental Forest (FEF) near Fraser, Colorado, USA. A *breakpoint* was defined by the flow where the fitted functions intersected and this was interpreted as the transition between phases of transport. There were markedly different statistical and sedimentological features associated with flows that were less than or greater than the breakpoint discharge. The fitted line for less-than-breakpoint flows had a lower slope with less variance due to the fact that bedload at these discharges consists primarily of small quantities of sand-sized materials. In contrast, the fitted line for flows greater than the breakpoint had a significantly steeper slope and more variability in transport rates due to the physical breakup of the armour layer, the availability of subsurface material, and subsequent changes in both the sizes and volumes of material in transport.

The model had good application to data from the eight FEF sites and in four supplementary channels to which the procedure was applied (Ryan *et al*., 2002). Results comparing the relative discharge for the breakpoint were similar in the studied channels, indicating that a substantial change in the nature of bedload transport occurs at about the same relative level of flow. Specifically, the relative breakpoint (R_{br}) is the breakpoint discharge (Q_{br}) expressed as a percentage of the bankfull discharge ($Q_{1\cdot5}$, a surrogate for the bankfull discharge):

$$R_{br} = 100(Q_{br}/Q_{1\cdot5})$$

The R_{br} was, on average, 80 per cent of the 1·5-year return interval flow for the 12 sites. Though the value ranged from about 60 to 100 per cent, there was no trend in R_{br} for channels with differing slope, size, or grain size characteristics. In the earlier analysis, we concluded that variation in R_{br} was more likely due to error associated with determining the breakpoint or bankfull discharge rather than morphologic variation between the channels.

These earlier results were derived from data collected in basins that, for the most part, have quite similar geology and runoff regimes. In this paper we evaluate further the R_{br} for channels with a wider range of geomorphic conditions. The piecewise regression model is applied to measured bedload data from 19 sites located in Colorado and Wyoming, including data from the eight FEF sites presented previously (Table I, a-h). Data from the four supplementary sites, discussed briefly in Ryan *et al*. (2002), are described more fully in this paper. These include the South Fork Cache la Poudre in Colorado, and Coon Creek, East Fork Encampment River, and Little Granite Creek in Wyoming. Added in the present series of comparisons are data from a fourth site in Wyoming (Cache Creek), two sites in the Arkansas River basin of Colorado (Halfmoon and Hayden Creeks), and four channels in the San Juan mountains of Colorado (Middle Fork Piedra River, East Fork San Juan, and one of its tributaries, Silver Creek, and the Florida River above Lemon Reservoir) (Table I).

Objectives

The primary objective of this study is to compare and contrast patterns of bedload transport observed in a range of coarse-grained channels and relate differences to channel and basin characteristics. First, we discuss observed differences (and similarities) in sediment loads for 19 gravel-bed channels and sources of variability in transport rates. Second, a piecewise regression analysis, described in Ryan *et al*. (2002), is used to objectively define a breakpoint and phases of transport in these channels. Third, we test for differences in the relative breakpoint in channels ranging from step–pool to almost-braided. Finally, we compare and contrast grain sizes moved at these sites under differing levels of flow.

Table I. Characteristics of 19 coarse-grained channels in Colorado and Wyoming

Reference letter	Site	Channel type	Management history	Years of gauge record	Years sediment collected	Drainage area (km^2)	Site elevation (m)	Mean slope (m m^{-1})	Primary lithology
a	St. Louis 1	Step–pool	Diversion	1934–present*	1992–93, 95–97	55·6	2790	0·039	Gneiss/granite
b	St. Louis 2	Plane	Diversion	1934–present*	1992–93, 95–97	54·2	2820	0·017	Gneiss/granite
c	St. Louis 3	Riffle	Diversion	1934–present*	1992–93, 95–97	54·0	2820	0·016	Gneiss/granite
d	St. Louis 4	Plane	Free-flowing, pristine	1943–present†	1992–93, 95–97	33·8	2910	0·019	Gneiss/granite
e	St. Louis 4a	Riffle	Free-flowing, pristine	1943–present†	1993, 95–97	33·5	2910	0·019	Gneiss/granite
f	St. Louis 5	Step–pool	Free-flowing, pristine	1943–present†	1992–93, 95–97	21·3	2925	0·048	Gneiss/granite
g	E. St. Louis Cr.	Step–pool	Free-flowing, pristine	1943–present	1993, 95–97	8·0	2905	0·050	Gneiss/schist
h	Fool Cr.	Plane	Free-flowing, harvested	1941–present	1993, 95–97	2·9	2910	0·044	Gneiss/schist
i	Little Granite Cr.	Plane	Free-flowing, pristine	1982–92	1982–1993, 97	54·8	1948	0·020	Sedimentary
j	S. Fk. Poudre R.	Riffle	Free-flowing, some private ownsership	1956–79	1989, 97	231·0	2408	0·007	Granite/gneiss
k	Coon Cr.	Plane	Free-flowing, harvest, tie-driven	1983–1997	1993, 95, 96	16·7	2906	0·031	Granite/gneiss
l	E. F. Encampment R.	Plane	Free-flowing, historical harvest, tie-driven	1983–1997	1993, 95, 96	9·1	2921	0·038	Granite/gneiss
m	Halfmoon Cr.	Riffle	Free-flowing, historical harvest	1946–present	1998, 99	61·1	2996	0·015	Granite/gneiss
n	Hayden Cr.	Plane	Free-flowing, harvested	No gauge	1998, 99	46·5	2240	0·025	Sedimentary
o	Cache Cr.	Plane	Free-flowing, pristine	1962–2002	1999	27·5	2057	0·021	Sedimentary
p	E. F. San Juan R.	Riffle–braided	Free-flowing, some private ownsership	1956–present	1999	166·0	2440	0·008	Volcanic
q	Silver Cr.	Plane	Free-flowing, pristine	No gauge	1999	15·7	2530	0·045	Volcanic
r	Upper Florida R.	Step–pool	Free-flowing, pristine	1973–present	1999	115·0	2550	0·013	Granite
s	M. Fork Piedra	Riffle	Free-flowing, pristine	No gauge	2001	86·0	2484	0·018	Volcanic/sed.

* Correlated with USGS gauge on St. Louis Creek # 09026500 (Ryan *et al.*, 2002)
† Correlated with USFS gauge on East St. Louis Creek

Study Sites

All of the study sites described herein are located on streams in Colorado and Wyoming where the US Forest Service or the US Geological Survey have initiated studies on sediment transport over the past 10–20 years. About half the data have been published previously (e.g. Leopold, 1992, 1994; Ryan and Troendle, 1996; Troendle *et al*., 1996, 2001; Ryan, 2001; Ryan and Emmett, 2002; Ryan *et al*., 2002). The streams are located within either the Middle or Southern Rocky Mountain provinces (Hunt, 1974) and are tributaries to the Snake, Platte, Colorado, Arkansas, and San Juan Rivers (Figure 1). Most sites are located near presently or historically operating gauging stations and represent a range of channel sizes and types, as outlined in Tables I and II. Drainage areas range from 2·9 to 230 km^2 and the sites are between 2000 to 3000 m in elevation. Channels are characterized by moderate to steep slopes (0·01 to 0·05 m m^{-1}). The median grain size of the channel surfaces ranges from very coarse gravel to coarse cobble sizes and a sorting index

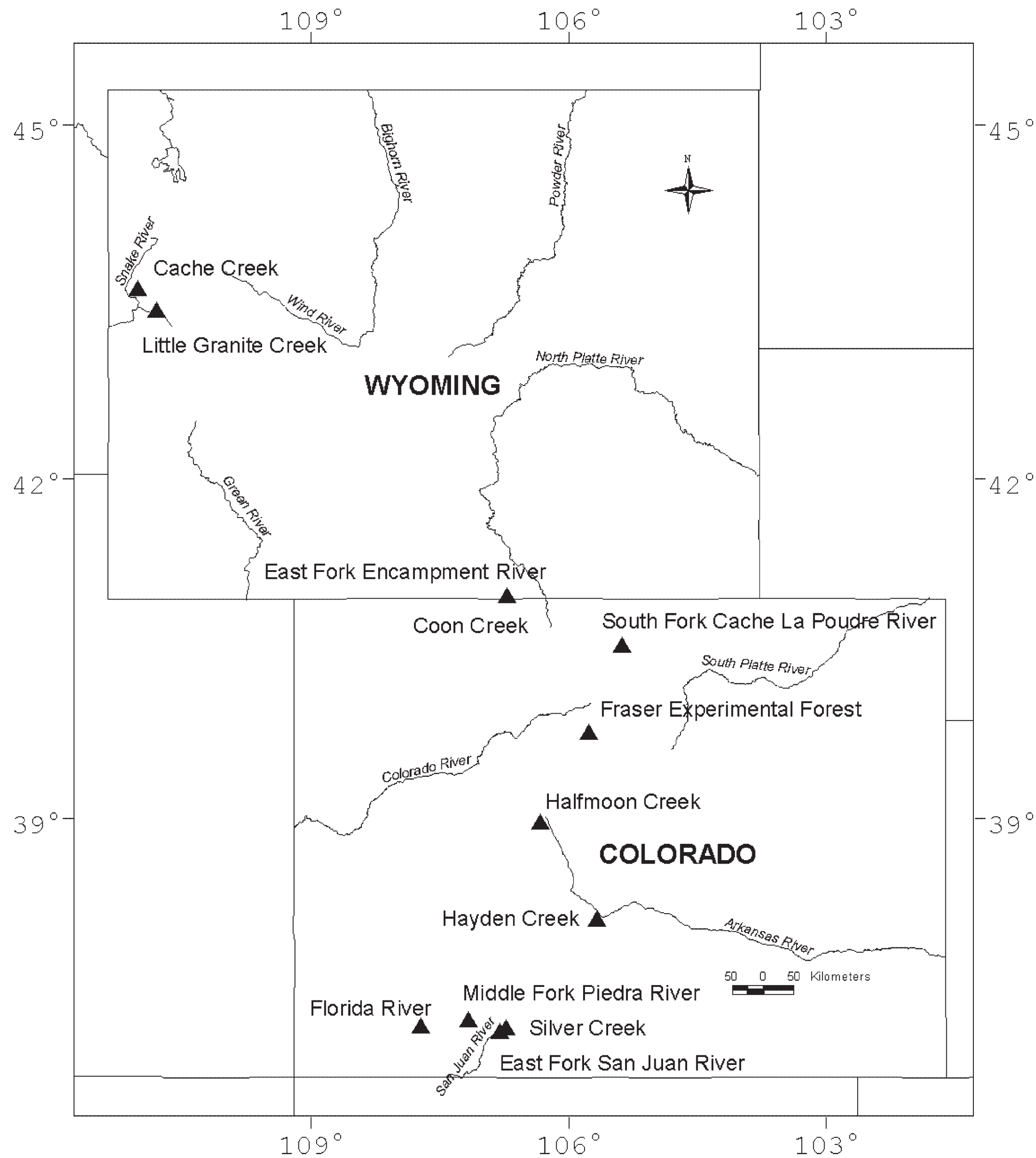

Figure 1. Location of bedload study sites and major rivers in Colorado and Wyoming. Eight of the sites are located at the Fraser Experimental Forest in the headwaters of the Colorado River basin.

Table II. Breakpoints and channel parameters for 19 coarse-grained channels in Colorado and Wyoming

Reference letter	Site	Discharge at breakpoint Q_{br} (m^3 s^{-1})	Standard error breakpoint estimate	95% lower conf. limit (m^3 s^{-1})	95% upper conf. limit (m^3 s^{-1})	Bankfull discharge $Q_{1\cdot5}$ (m^3 s^{-1})	Relative breakpoint ($Q_{br}/Q_{1\cdot5}$) (%)	Bankfull width w (m)	Bankfull depth H (m)	Grain size D_{16} (mm)	Grain size D_{50} (mm)	Grain size D_{84} (mm)	Sorting index*	D_{84}/H_{bf} (m m^{-1})
a	St. Louis 1	3·41	0·56	2·31	4·52	4·41	77	6·8	0·47	27	129	362	3·8	0·77
b	St. Louis 2	4·50	0·33	3·86	5·13	4·75	95	7·6	0·43	34	76	172	2·2	0·40
c	St. Louis 3	4·04	0·65	2·76	5·31	4·59	88	9·3	0·36	34	82	181	2·3	0·50
d	St. Louis 4	2·07	0·29	1·49	2·64	3·61	57	7·0	0·38	36	91	185	2·3	0·49
e	St. Louis 4a	2·84	0·18	2·49	3·19	3·37	84	8·2	0·34	23	79	167	2·8	0·49
f	St. Louis 5	2·35	0·13	2·10	2·61	2·63	89	5·3	0·38	30	146	543	4·3	1·43
g	E. St. Louis Cr.	0·76	0·06	0·63	0·88	0·86	88	3·0	0·30	14	51	142	3·2	0·47
h	Fool Cr.	0·22	0·02	0·18	0·26	0·30	73	2·0	0·17	11	38	100	3·0	0·59
i	Little Granite Cr.	4·83	1·19	2·50	7·16	5·95	81	9·6	0·45	24	89	207	3·0	0·46
j	S. Fk. Poudre R.	8·57	0·43	7·72	9·41	11·44	75	13·2	0·59	32	70	115	1·9	0·19
k	Coon Cr.	2·62	0·15	2·32	2·93	2·77	95	5·9	0·33	26	83	214	2·9	0·65
l	E. F. Encampment R.	1·50	0·15	1·24	1·84	1·45	103	6·2	0·25	20	49	135	2·6	0·54
m	Halfmoon Cr.	2·51	1·80	0·00	6·04	6·23	40	8·6	0·52	34	62	110	1·8	0·21
n	Hayden Cr.	1·68	0·21	1·28	2·08	1·92	88	5·4	0·27	20	68	140	2·7	0·52
o	Cache Cr.	2·02	0·19	1·65	2·38	1·84	110	5·2	0·40	11	46	115	3·3	0·29
p	E. F. San Juan R.	10·23	0·89	8·48	11·97	15·66	65	17·2	0·53	21	50	112	2·3	0·21
q	Silver Cr.	0·98	0·09	0·81	1·16	1·30	75	4·2	0·30	12	31	71	2·4	0·24
r	Upper Florida R.	10·66	0·43	9·81	11·51	14·50	74	12·0	0·77	35	218	550	4·4	0·71
s	M. F. Piedra R.	7·59	0·36	6·88	8·29	10·05	76	13·5	0·39	20	79	210	3·3	0·54

* This index is as defined by Andrews (1983) where the D_{value} is a percentile on the cumulative-percentage-finer-than scale: $\sigma = \frac{1}{2}\left(\frac{D_{84}}{D_{50}}\right) + \left(\frac{D_{50}}{D_{16}}\right)$

ranges from 2 to 4 (poorly to very poorly sorted) (Table II). Channel types include step–pool, plane-bed, pool–riffle, and one moderately braided site (East Fork San Juan).

The lithology of the studied areas is quite diverse (Tweto, 1979; Love and Christiansen, 1985) (Table I). Most of the sites are underlain by granite, gneiss, and schist bedrock while other streams flow through areas with sedimentary formations including sandstone, conglomerate, and limestone. The East Fork San Juan and Silver Creek are underlain by complex volcanic formations consisting of layered lava flows. Nearly all of the sites have been glaciated and are mantled by till and glacial outwash correlative of Bull Lake and Pinedale aged deposits (e.g. Nelson and Shroba, 1998). At most sites, hillslopes are relatively stable under the current climatic regime and vegetative cover, though there is evidence of active mass wasting in the past (Caine, 1986). Exceptions include Little Granite Creek, where there are active, deep-seated earthflows, and the sites underlain by volcanic formations where debris chutes and ravel from unstable terrain deliver materials directly to the stream system (Ryan, 2001).

While the studied sites are hydrologically heterogeneous in terms of the timing of the annual peak flow, there are common patterns that reflect the influence of snowmelt on the annual hydrograph (Jarrett, 1990). Peak discharges occur typically between May and July and low flow occurs in January or February. A second peak discharge frequently occurs in late summer at the sites in the San Juan mountains due to the influence of the Arizona monsoon. Most of the study sites are free-flowing, though three of the FEF sites are located in channels from which an average of 40 per cent of the total annual flow is diverted (Ryan, 1994). However, water is rarely diverted from these sites at high flow when bedload is usually collected. Small portions of flow at East Fork San Juan and South Fork Cache la Poudre sites are diverted for household or agricultural purposes upstream of the study sites. Some of the watersheds have been deforested to different degrees (e.g. Troendle and King, 1985; Troendle *et al.*, 2001) and a few of the channels were tie-driven in the historical past (Young *et al.*, 1994). Other areas have been selectively cut, to an extent which is either relatively minor or unknown.

Methods

Samples of bedload were usually collected using a 3×3 inch (76×76 mm) Helley–Smith bedload sampler (Helley and Smith, 1971) either while wading or from temporary sampling platforms (Martinez and Ryan, 2000). Samples from Silver Creek, East Fork San Juan, Middle Fork Piedra, and the Florida River were collected using a wading version of an Elwha bedload sampler with a 4×8 inch (102×203 mm) (Childers *et al.*, 2000; Ryan, 2001); the upper size limit of this sampler is about 200 mm. By comparison, the upper size limit of the Helley–Smith sampler is about 76 mm, which potentially truncates the upper end of the bedload distribution. However, analyses of sediment obtained from settling ponds at several sites in the region indicate that 85–90 per cent (by weight) of the grains trapped in the ponds would pass easily through the opening of a Helley–Smith sampler (Wilcox *et al.*, 1996; Ryan and Porth, 1999), suggesting that the truncation is relatively minor. Regardless of the sampler deployed, all measurements made for a site use the same type of device.

Bedload was sampled once or twice daily over several weeks during snowmelt runoff using the Single Equal Width Increment (SEWI) method (Edwards and Glysson, 1998). Bedload samples were dried and sieved using standard sedimentological methods (Folk, 1968). Full phi-interval sieves, ranging from 0·5 to 64 mm, were used to separate samples into grain size classes. Rate of transport was calculated using the total weight of the sample (in kg) divided by the total sampling time (number of verticals and time in seconds) and width of the sampler (in m) to obtain the unit bedload transport rate (kg m^{-1} s^{-1}). This value is multiplied by the width of the channel (in m) to obtain the mean transport rate through the channel cross-section (in kg s^{-1}). Fractional transport rates were calculated by dividing the total transport rate by the proportion of sample retained in each sieved phi class. Flow was estimated for each bedload observation using either the mean discharge observed for the sampling period from the stage–discharge relationship or by a direct discharge measurement made immediately following bedload collection. Discharge was determined using standard methods (Buchanan and Somers, 1969; Nolan and Shields, 2000). Velocity was measured with Price AA or pygmy current meters. Channel width, mean depth, and mean velocity were measured in association with discharge measurements (Table II). Water surface slope, approximating the energy gradient, was surveyed in the immediate vicinity of the sampling site over a range of discharges, including bankfull. The particle size distribution of the bed surface was assessed from grid-based pebble counts (Wolman, 1954) using between 100 and 400 particles. Samples were collected from the subsurface at most of the sites using a barrel method (e.g. Milhous *et al.*, 1995) with a modified 55-gallon barrel.

Flow frequencies were calculated using Log Pearson type III analysis on the annual maximum series for the period of record (Table I), whenever data were available (US Interagency Advisory Committee on Water Data, 1982). At the few ungauged sites, on-site measurements of discharge were correlated with nearby gauges to estimate the return frequency for flows. There is generally a good relationship between the discharge with a 1·5-year return frequency ($Q_{1.5}$)

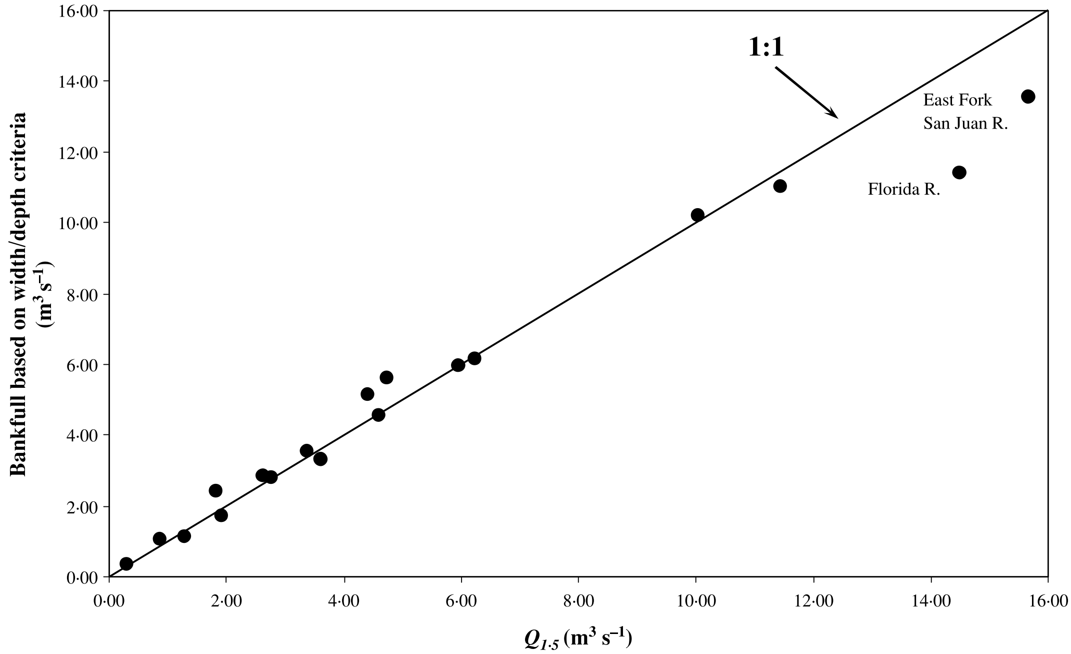

Figure 2. Comparison of two estimates used to estimate bankfull discharge. The *x*-axis is the 1·5-year flow determined from Log Pearson III analysis. The *y*-axis is an estimate based on the discharge where the width:depth ratio is a minimum.

and the flow that just fills the banks in most of our channels (Troendle *et al*., 1996). As a check on this relationship, measured width:mean depth ratios were plotted against measured flow to determine the discharge at which the width: depth ratio is a minimum. Based on measurements of hydraulic geometry, this value is a useful indicator of bankfull discharge (e.g. Carling, 1988). The two estimates of bankfull were generally in agreement and there were no consistent differences between the 1·5-year flow and the minimum width:depth ratio that would imply a bias towards one method (Figure 2). On average, $Q_{1\cdot5}$ was slightly (though not significantly) less than the width:depth method ($0\cdot4 \pm 12\cdot6$ per cent – one standard deviation). For the sake of consistency and because there is less subjectivity in the estimate, we use $Q_{1\cdot5}$ as determined from the Log Pearson analysis for the estimate of bankfull discharge, whenever possible.

Bedload rating curves

Several models were used to relate measured rates of bedload transport to flow, including power and piecewise regression models. The power function ($E[y] = ax^b$) (where x is discharge (in $m^3\ s^{-1}$) and y is rate of bedload transport (in $kg\ s^{-1}$)) provided a good first approximation of the fit over the full range of measured discharges. In most cases, the fit accounted for as much variance as the piecewise regression applied to the same dataset (Ryan and Emmett, 2002; Ryan *et al*., 2002). However, there were often discontinuities in the relationship between flow and sediment and a single power function was not always appropriate for the full range of data. Frequently, the exponents (b) for flows less than about 50 per cent of bankfull were not statistically different from 1, indicating a linear relationship for this range of flows. The exponent of the power function typically increased as data from higher flows were included in the fit. Hence, the exponent of a power model appears to be related, in part, to the range of data used in the fitting procedure.

Piecewise regression implicitly recognizes different functions over varying ranges of flow (Neter *et al*., 1989). The form of the model used here fits linear segments to different ranges of data, though other types of functions may be used. Breakpoints are values on the x-axis where a change in the slope of the different linear relationships was defined. We assume that, given what is understood about the nature of bedload movement in natural channels, the function should be continuous at all points, including the breakpoint. When there is only one breakpoint (at $x = c$) the model can be written as:

$$E[y] = a_1 + b_1 x \quad \text{for } x \le c$$
$$E[y] = a_2 + b_2 x \quad \text{for } x > c$$

In order for the function to be continuous, the two equations for $E[y]$ need to be equal when $x = c$,

$$a_1 + b_1 c = a_2 + b_2 c \Rightarrow a_2 = a_1 + c(b_1 - b_2)$$

The function can be rewritten as follows, substituting for a_2:

$$E[y] = a_1 + b_1 x \quad \text{for } x \le c$$
$$E[y] = \{a_1 + c(b_1 - b_2)\} + b_2 x \quad \text{for } x > c$$

The model was applied using several fitting procedures (Ryan *et al*., 2002). The estimated equations from these procedures were usually identical and so only the results using generalized least squares (GLS) are presented.

The reliability of the breakpoint estimate depends on several factors, including a dataset containing a substantial number of observations collected over a range of discharges. Problems were encountered on datasets where samples were measured primarily at discharges less than bankfull (e.g. Hayden Creek). One common difficulty when fitting any regression model to bedload data is the non-normality and heterogeneous variance of the model results (Ryan and Porth, 1999; Ryan *et al*., 2002). When this occurs, the standard errors of the estimates for each of the parameters are suspect. Therefore, a bootstrapping procedure was used to obtain non-parametric estimates of the standard errors and confidence intervals for model parameters (Efron and Tibshirani, 1993). Bootstrapping involves resampling from the original dataset with replacement in order to obtain a secondary dataset. The model is then fitted to the secondary dataset and the parameters from the fitted equations (the bootstrap estimates) are retained. The procedure is repeated a specified number of times and standard errors for the parameters are calculated as the standard deviation of the bootstrap estimates (Efron and Tibshirani, 1993). We used 1000 iterations, meaning 1000 secondary datasets were generated, to obtain confidence limits on the bootstrap estimates.

Results

Sediment loads

The amount of material moved as bedload at a site is influenced by a number of factors, including runoff patterns, volume and caliber of the sediment supply, and changes in land use. In this study, mean rates of bedload measured at bankfull discharges ranged from almost 5 kg s^{-1} at the East Fork San Juan site to 0·004 kg s^{-1} at Cache Creek. Within the same watershed, sediment loads typically increase moving downstream into increasingly larger channels. For example, the relationship between drainage area and the rate of transport at bankfull discharge for the St. Louis Creek watershed was relatively strong (Figure 3). There is about an order of magnitude increase in the rate of transport from small watersheds (drainage area about 5 km^2) to lower St. Louis Creek (about 50 km^2). When normalized by drainage area, the rate of sediment transport per basin area at bankfull is comparable for sites within this watershed as well as areas with similar geologic settings. For instance, in several areas underlain by granite, gneiss, and schists, the mean normalized rate of transport was generally between 0·001 and 0·002 kg s^{-1} km^{-2} (Figure 4). In contrast, the mean was on the order of 0·025 kg s^{-1} km^{-2} for areas underlain by unstable volcanic formations or approximately an order of magnitude greater than that observed for the granitic/metamorphic settings. Nevertheless, geologic setting alone does not entirely explain the differences in rates of transport. Normalized rates at Halfmoon Creek (granitic) were about twice the values observed in similar settings, perhaps reflecting a land use influence. Conversely, normalized rates were very low (0·0002 kg s^{-1} km^{-2}) for the Florida River site (also granitic), which is indicative of relatively low supplies of sediment in this partially bedrock system (Ryan, 2001). Normalized rates observed for areas underlain primarily by sedimentary formations ranged from 0·0001 at Cache Creek to 0·0025 at Little Granite Creek, likely reflecting differences in supply mechanisms within this broad classification of rock type.

Hysteresis and transport variability

In previous research on gravel transport, the presence of annual (or seasonal) hysteresis has been suggested to explain some of the observed variation in rates of bedload transport (e.g. Klingeman and Emmett, 1982; Reid *et al*., 1985). In some cases, clockwise hysteresis has been observed, where higher transport rates are measured on the rising limb of the hydrograph compared to those measured on the falling limb. This is attributed to a greater supply of sediment in-channel prior to the onset of runoff and an exhaustion of sediment over the duration of high flow (Nanson, 1974; Dunne and Leopold, 1978; Ferguson, 1987; Moog and Whiting, 1998). A counterclockwise pattern happens when there are higher rates of transport on the falling limb of the hydrograph. This pattern can occur when the coarse surface layer is disrupted at peak flows and sand and gravel from the subsurface are available for transport on the

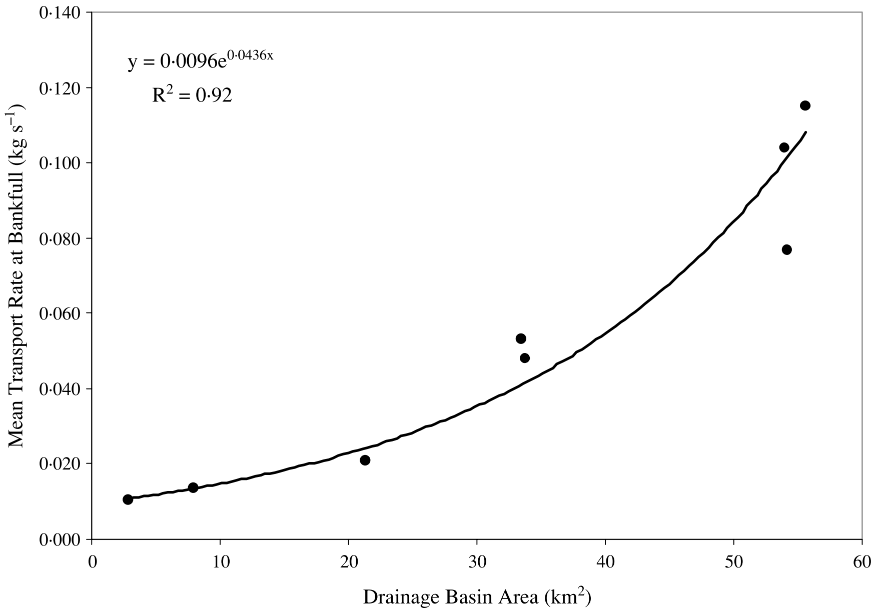

Figure 3. Predictive relationship between transport rate at bankfull and drainage area for main stem and tributaries to St. Louis Creek, Fraser Experimental Forest.

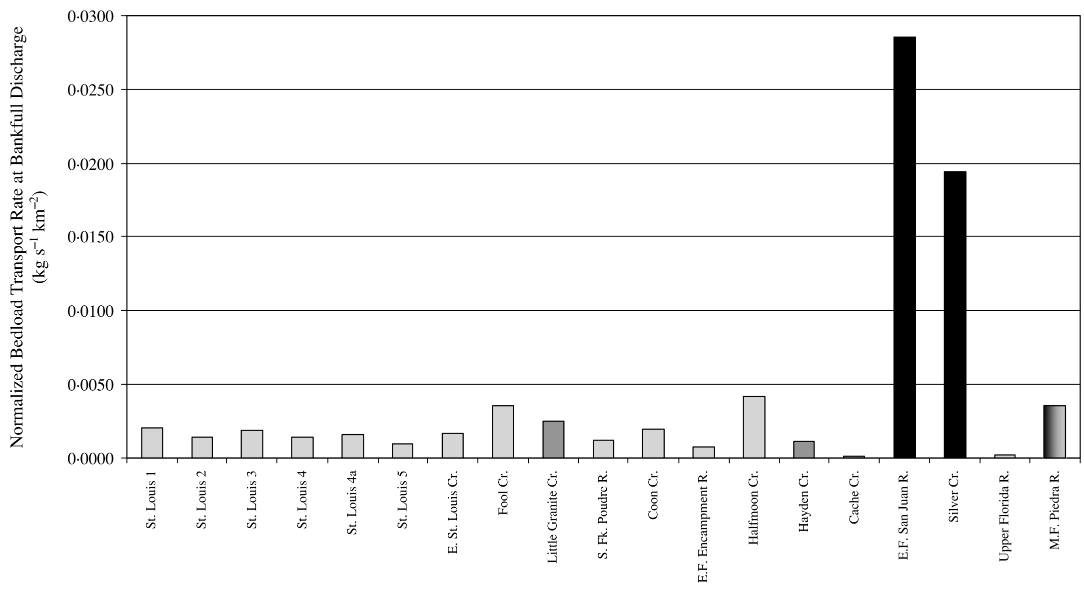

Figure 4. Normalized rates of bedload transport measured at bankfull discharges at 19 sites. The data have been normalized by dividing by basin area. Areas underlain primarily by granite, gneiss, and schist are shown in light grey. Areas underlain primarily by sedimentary bedrock are shown in medium grey. Areas underlain by extrusive volcanic materials are shown in dark grey. The Middle Fork Piedra R. shows a gradation because it lies on a transition between volcanic material and sedimentary bedrock.

recessional limb, at rates higher than observed before the peak (e.g. O'Leary and Beschta, 1981; Klingeman and Emmett, 1982). This can also occur through episodic bank erosion, wood jam breakups, or mass wasting delivering sediment to the system on the waning limb of the hydrograph.

While some larger rates of transport were measured at relatively low discharges at several of our sites, not all were exclusive to the rising *or* falling limb. For instance, relatively high rates of transport were measured (though sporadically) during the rising limb of snowmelt runoff on Silver Creek (open circles in Figure 5q), suggesting a

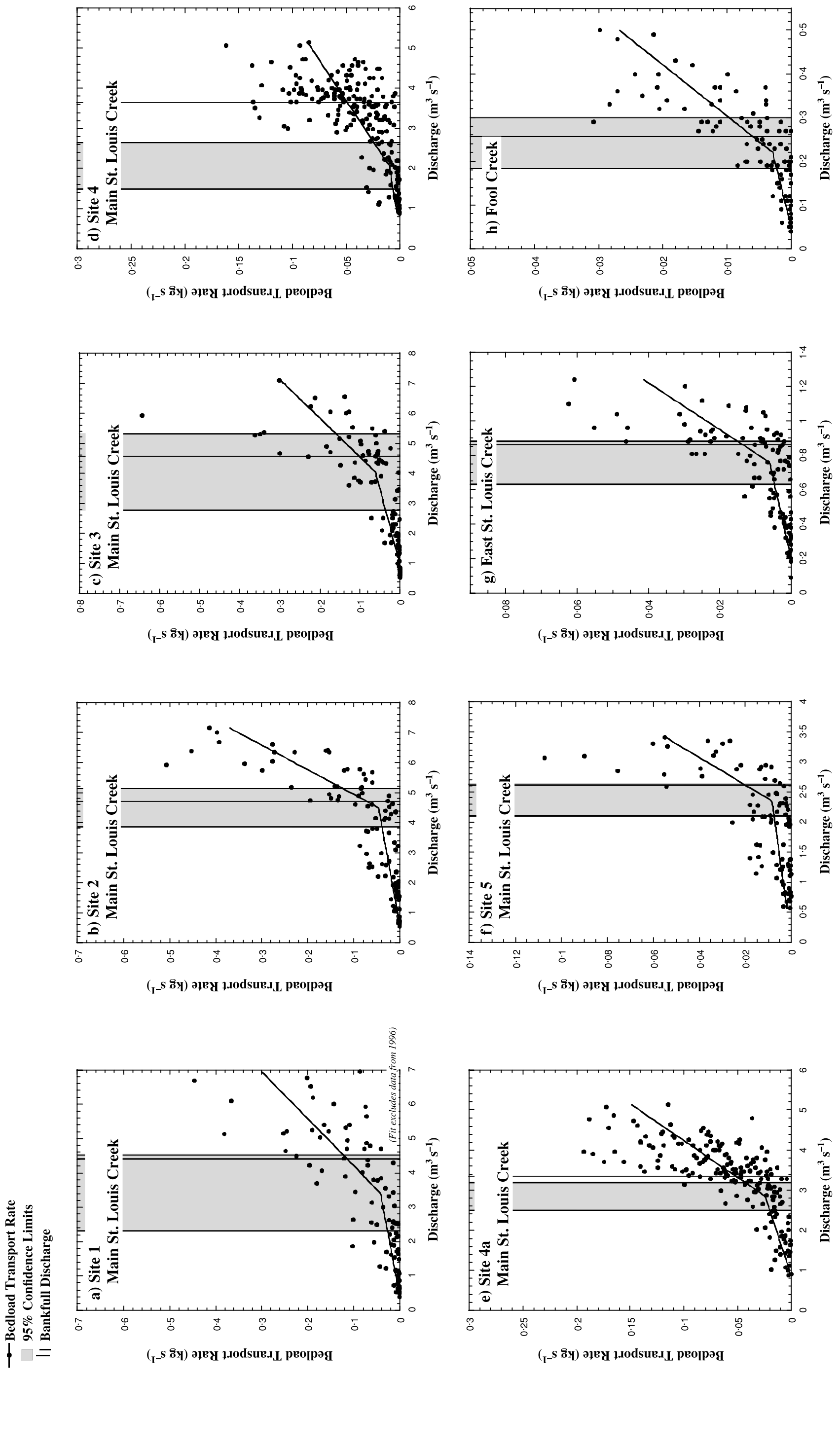

Figure 5. Sediment rating curves for 19 sites determined using piecewise regression analysis (thicker solid line). The thinner, vertical line is the estimate for bankfull discharge. The shaded area indicates the 95 per cent confidence limit for the breakpoint estimate.

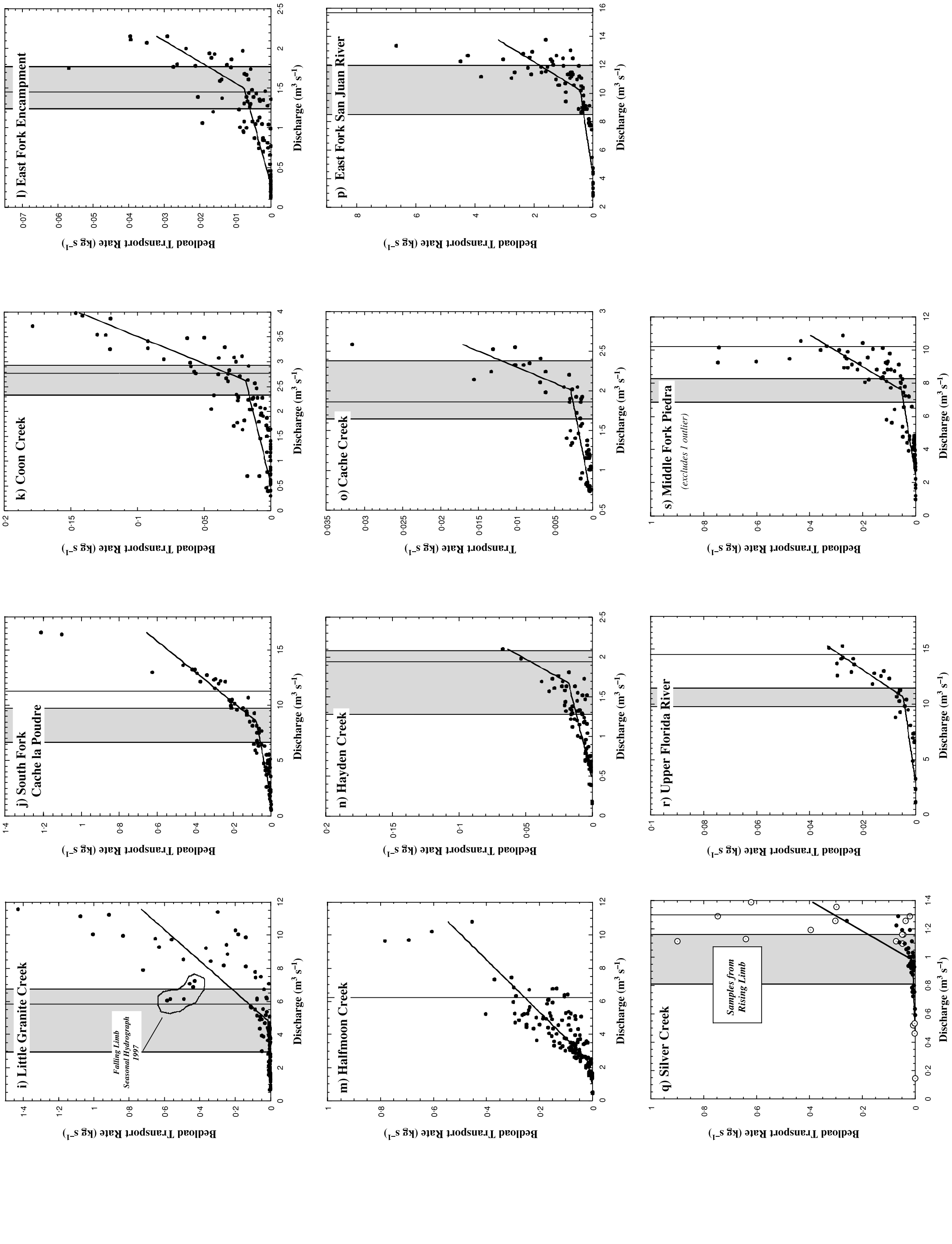

Figure 5. *(Continued)*

clockwise hysteresis effect. This was due largely to the mobilization of loosely bound slugs of sediment from avalanche chutes and debris slides that had been delivered directly to the channel upstream (Ryan, 2001). Lower rates of transport at comparable discharges on the recessional limb likely reflect a diminishing sediment supply. Clockwise hysteresis was also observed in the Coon Creek data, but only for one out of three years. A clockwise pattern also occurred at St. Louis Creek 1 in 1996 when low rates of transport were measured at high discharges. Counterclockwise hysteresis was observed at Little Granite Creek in 1997 when several large, coarse bedload samples were measured as flow began to wane (encircled on Figure 5i). These likely represent gravel from the bed and banks that continued in motion as the bed surface re-stabilized. Similar patterns were observed at Site 1 in St. Louis Creek in 1992 when relatively large amounts of sediment were trapped on the falling limb of the hydrograph (Ryan, 1994). The samples consisted primarily of sand and small gravel, which might support the idea that they were grains made available upon the breakup of the surface layer. However, because flows that year were hardly sufficient for widespread disruption of the surface, it is more likely that the material originated from finer patches upstream.

These observed patterns of hysteresis contribute to the overall variability of rates of sediment transport. However, there was no consistent pattern observed for the 19 sites that would suggest a universal process, indicating that changes in the seasonal supply of sediment are specific to conditions at individual sites. Moreover, we found that the hysteric patterns and the accompanying variability in measurements of bedload did not detract substantially from the piecewise regression model's ability to predict breakpoints. For instance, a breakdown of the measurements from both the rising limb and falling limbs on the Silver Creek data (Figure 5q) indicated that while the slope of the greater-than-breakpoint flows was steeper for data from the rising limb, the position of the breakpoint was within the confidence limits determined using values from the falling limb of the hydrograph.

Initiation of significant gravel transport – breakpoint analysis

The piecewise regression model was used to generate a breakpoint with two linear functions for each of the 19 sites (Figure 5). In all cases, the intercept (a_1) was negative. This occurs because bedload movement becomes undetectable (or goes to near zero) before flow becomes zero. However, because a negative transport rate has little physical meaning, the predicted rate of transport for very low discharges would be zero. In some instances, the intercept was not significantly different from zero, but was left in the model because the line better fitted the bulk of the data at lower discharges. While the model usually converged to produce a breakpoint, one needs to evaluate the confidence limits of the parameters and the variance accounted for by the model to determine if it is an appropriate fit. In some cases, the breakpoint was well-defined with relatively narrow confidence limits (e.g. Figure 5e and s). In other cases, the confidence bands were relatively wide (Figure 5i and n). In one case, the model converged, but the confidence bands encompassed almost the full range of data (Figure 5m, bands not shown) indicating that the breakpoint is suspect. Here, other information, such as the presence of coarse grains in the samples, may be needed to define whether a change in 'phase' has occurred. More generally though, the piecewise regression fit was useful for isolating two distinct phases of bedload transport from which the breakpoint discharge could be determined. The fitted lines for flows that are less than the breakpoint have a lower slope (median value = 0·012) than the slope of the fitted lines for flows greater than the breakpoint, which were between 5 and 25 times greater (median slope = 0·08).

Similar to our earlier results, the breakpoint occurred at flows which were about 80 per cent of the bankfull discharge (R_{br}), with one standard deviation ranging between 65 and 95 per cent (Figure 6, Table II). This supports the idea that a substantial change in the nature of sediment transport begins at flows approaching bankfull. There were, however, some sites where the R_{br} was outside of the 65 to 95 per cent range. For instance, the breakpoint was clearly low at Halfmoon Creek where the calculated R_{br} was 40 per cent. Grain size data for this site (Figure 7b) suggest that the onset of coarse gravel transport occurs between 40 and 50 per cent of bankfull, supporting the results of the piecewise regression and indicating that there is transport of larger grains at relatively low discharges. Hence, we are reasonably confident in our assertion that the R_{br} is comparatively low in this system. The R_{br} for Cache Creek was 110 per cent which is substantially higher than that determined for other sites. This is likely due to a lack of correspondence between bankfull and the 1·5-year flow at this site, which affects the value of R_{br}. However, for the sake of consistency, we continue to use this estimate in the following series of comparisons.

Ryan *et al*. (2002) observed no significant trend between R_{br} and a number of measured basin and channel parameters for the original 12 sites. Specifically, water surface slope, median grain size of the channel bed (D_{50}), dimensionless grain size (D_{50}/w) where w is the channel width, drainage area, relative roughness ($D_{84}/H_{1\cdot5}$) where $H_{1\cdot5}$ is the mean depth at bankfull, and a sorting index for the channel surface (Andrews, 1983) were used in these earlier comparisons. Regression analysis was repeated using the expanded dataset and, similar to the earlier findings, none of the trends were statistically significant. Conceivably, there may be other factors that cause fluctuations in R_{br}, such as sediment supply (Dietrich *et al*., 1989) or bed compaction (Church *et al*., 1998), that we are unable to directly test for at this time.

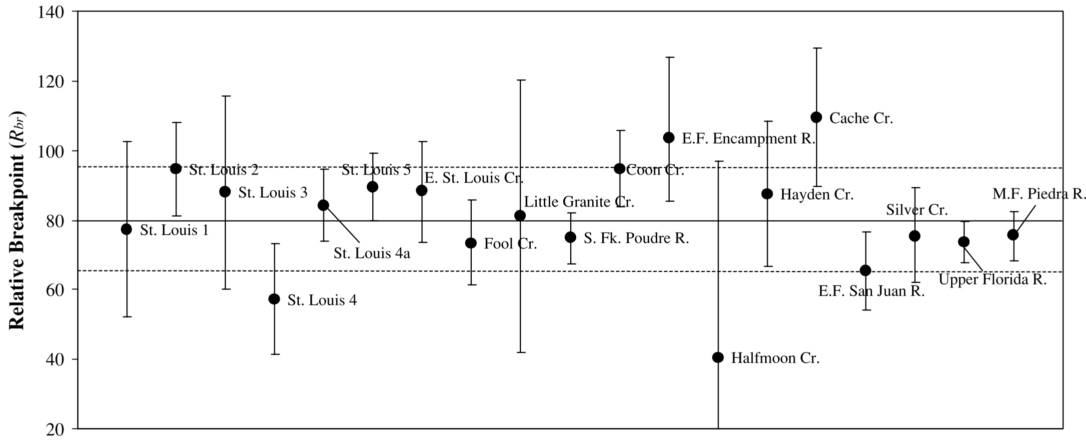

Figure 6. Range of relative breakpoints (R_{br}) and 95 per cent confidence limits determined for each of 19 gravel-bed channels. Solid horizontal line is the mean and the dotted lines are the 95 per cent confidence limits of the relative breakpoint for all sites.

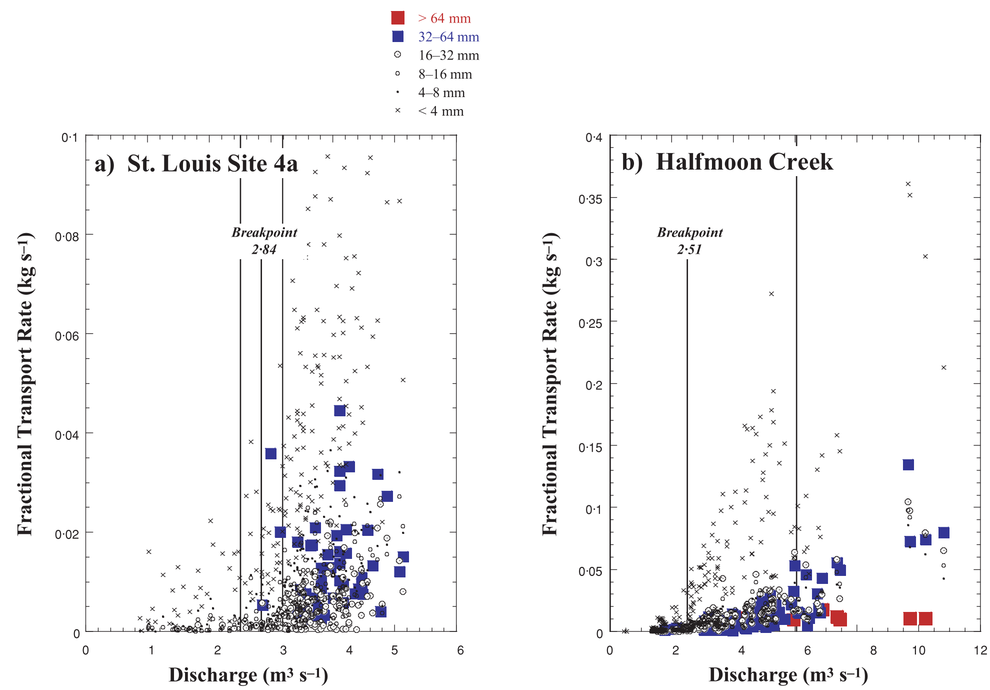

Figure 7. Examples of rates of transport for individual size classes. The breakpoint value and confidence limits on this estimate are depicted by vertical black lines. Note that the rate of transport for fines (<4 mm) is typically greater than that for coarse size fractions, a pattern common in gravel-bed channels. Values of zero are removed for illustration purposes.

We then evaluated the hypothesis that R_{br} would differ by channel type, which, in a sense, tests the combined influence of a number of parameters (for example, step–pool channels tend to be steeper, narrower, and rougher than either plane-bed or pool–riffle channels). The sites were separated into classes using channel types identified in the field and an analysis of variance (ANOVA) was performed to determine whether there were significant differences between the three groups. The results of Levene's test (Milliken and Johnson, 1984) indicated variance between groups was homogeneous ($p = 0{\cdot}291$). A one-way ANOVA indicated no significant differences in R_{br} between the groups ($F_{2,16} = 1{\cdot}679$, $p = 0{\cdot}218$). The mean for group 1 (step–pools) was 80·8 (71·5–90·1), the mean for group 2 (plane-bed) was 86·3 (73·8–98·9), and the mean for group 3 (pool–riffle) was 70·6 (46·9–94·2); the numbers in

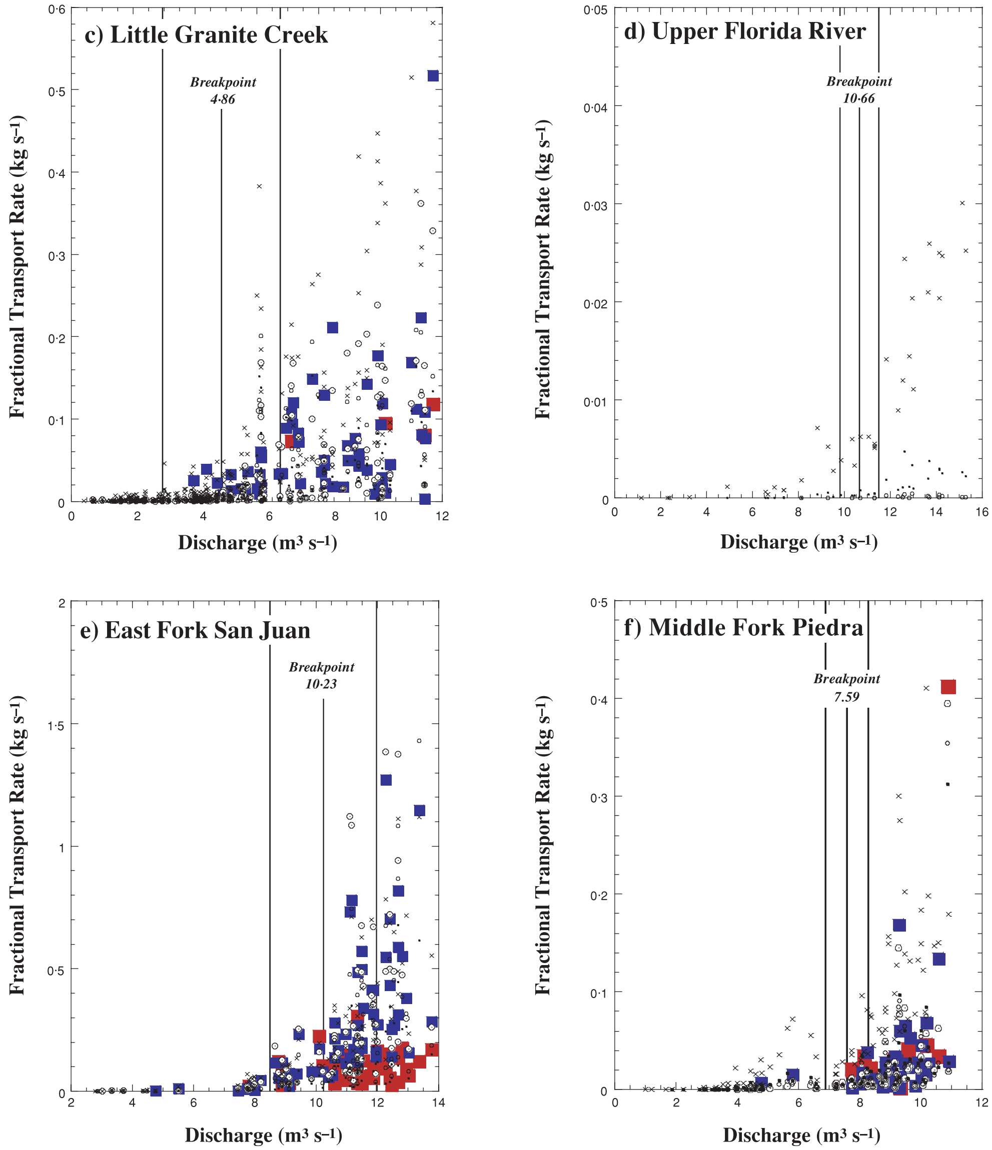

Figure 7. (*Continued*)

parenthesis are the 95 per cent confidence limits on the estimate of the mean. From this series of tests, we conclude that there were no statistically significant differences in the range of R_{br} for the types of channels considered in this study.

Rates of fractional transport at varying discharge

Rates of transport were determined for several grain size classes, then related to discharge to determine (1) whether there are grain sizes trapped at greater-than-breakpoint flows that are largely absent from samples collected at lower discharges; and (2) whether there is similarity in the grain sizes moved at the same relative level of flow. Samples from flows less than the breakpoint typically consist of sand with minor contributions of fine to medium sized gravel. Fine gravel often behaves similarly to sand in these systems and together they are the principal constituents of phase I transport (Ryan *et al*., 2002). The source of this material is most likely from more mobile patches and interstitial areas and does not signify widespread entrainment of the channel surface (Lisle, 1995). The primary constituent of

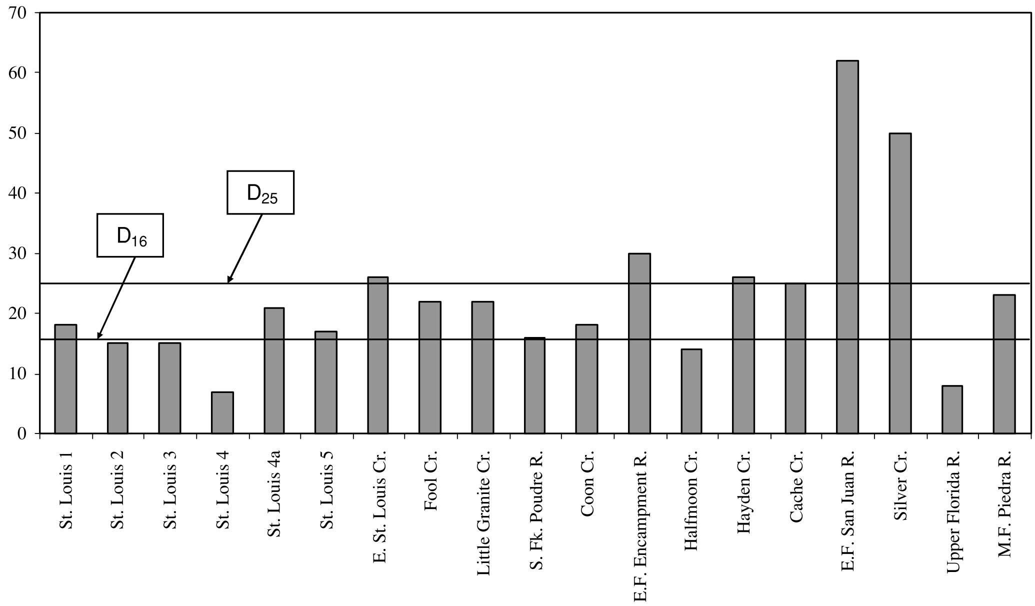

Figure 8. The breakpoint grain or the percentile of the bed surface that is trapped in samplers beginning at the breakpoint discharge. The breakpoint grain is identified from plots of fractional transport rates (Figure 7) and the percentile is determined from the particle size distribution of the channel surface. This value typically varies from D_{16} to D_{25} of the channel surface, though a few sites fall outside of this range.

bedload at high flow also tends to be sand and very fine gravel (Leopold, 1992), as indicated by higher rate of transport of fines (<4 mm) relative to coarser grains shown on most plots of fractional transport rates (Figure 7). Coarser particles are largely absent from samples until about the breakpoint discharge or within the confidence bands of the estimate, hence their movement during lower flow is considered to be essentially negligible. The smallest grain size class inclusive to samples at the breakpoint that is absent from low flow samples is termed the *breakpoint grain*.

Though the composition of bedload from different gravel-bed streams may be dissimilar, the breakpoint grain at most of our sites was usually coarse to very coarse gravel (16 or 32 mm). In this sense, the presence of 16 to 32 mm grains in bedload samples signifies the early stage of armour breakup (or the mobilization of the lower end of the 'framework' gravels, in the sense of Church *et al*., 1987). Moreover, there is similarity in the relative size of the breakpoint grain when scaled by the surface bed material. In Ryan *et al*. (2002), we observed that the breakpoint grain (formerly called 'grain size of interest') represents approximately D_{16} of the channel surface, the range being between D_7 and D_{26}. Here, using data from additional sites, we observed that the breakpoint grain is typically between D_{16} and D_{25} (Figure 8). Given the inherent difficulties in determining the distribution of gravel-bed surfaces from pebble counts (Marcus *et al*., 1995; Wohl *et al*., 1996) the results are remarkably consistent.

At three of the sites, however, the breakpoint grain was either considerably smaller or coarser than 16 or 32 mm. The breakpoint grain at the Florida River site was 8 mm (representing D_8 of channel surface; Figure 8). No grains larger than 16 mm were caught in any of the samples at high flows (Figure 7d), even though a sampler with a larger opening was used at this site. This probably reflects the fact that the channel bed is coarse and very well-armoured (or bedrock in some areas) with limited supplies of sediment from upstream. It is likely that only a minor portion of the bed is in motion at even higher discharges. Hence, the Florida River site may represent a more stable end member on a continuum of gravel-bed channels (Montgomery and Buffington, 1997). On the other end of the spectrum, coarse to very coarse gravel was trapped at all but the lowest flows on the East Fork San Juan (Figure 7e). Grains 64 mm and larger, representing D_{60} of the channel surface, were trapped beginning at the breakpoint; similar observations were made for Silver Creek (Figure 8). Hence, there seems to be a difference in the size of grains transported over a range of flows for these sites, likely representing the other end of the gravel-bed continuum (Montgomery and Buffington, 1997; Montgomery *et al*., 1999) and a comparatively high rate of sediment supply (Ryan, 2001). In addition, it appears that the overall grain size distribution of the bedload is more like that of the subsurface and surface at these two sites. This is discussed further in the following section.

Surface, subsurface, and bedload grain size distributions

For most sites, the composition of bedload is considerably finer than that of the channel surface or subsurface. Although the grain size distribution becomes increasingly similar to that of the subsurface at higher flows, the observed distributions are typically finer at flows up to 1·5–2 times the bankfull discharge (Figure 9a). However, given that there are restrictions on our ability to characterize the transport of coarser grains because of the size limitations of a 3-inch (76 mm) Helley–Smith sampler, the distributions at higher flows could be artificially fine for sites where this sampler was used. To investigate further the relationships between bedload, subsurface, and surface particle sizes, we compare sites where the Elwha sampler was deployed (East Fork San Juan, Silver Creek, and Middle Fork Piedra) against a general pattern observed with data from the Helley–Smith (Figure 9a).

Sediment loads for both the East Fork San Juan and Silver Creek are relatively coarse compared to the grain size distributions for the channel, even at moderate discharges. The bedload distribution at the East Fork San Juan becomes coarser than the subsurface beginning at about 50 per cent of bankfull and approaches that of the surface at flows near bankfull (Figure 9c). Comparable patterns were observed at Silver Creek, though they are more irregular (Figure 9b) because of probable hysteresis. In contrast, the grain size distribution for bedload on the Middle Fork Piedra is finer than that of the subsurface at flows near bankfull (Figure 9d), similar to our observations from sites where bedload was collected using a sampler with a smaller opening (Figure 9a). In these cases, the curves tend to be upwardly concave and steeper in the sand to small gravel range at low flows and become straighter and shift to the right with increasing discharge as more coarse material is incorporated into the sample. This occurred even when the size limitations were relaxed using a sampler with a larger orifice. Hence, because the patterns from the Middle Fork are similar to those observed for most of our other sites, the differences observed for the East Fork San Juan and Silver Creek more likely reflect variation in stream condition rather than simply the deployment of a larger sampler. Because eroding hillslopes and braided channels were observed elsewhere in the watershed (Ryan, 2001), we surmise that the relative coarseness of the bedload at these two sites is likely due to higher rates of sediment supply to the system.

Discussion and Conclusions

The goal of this investigation was to evaluate the dynamics of coarse sediment transported in a number of mountain streams and relate patterns of sediment transport to basin and channel characteristics. Each of the studied channels had its own 'bedload signal' in that the rate and size of materials transported are determined by the nature of flow and sediment particular to that system. Based on qualitative observations of erosion and sedimentation in the field, broad differences in sediment supply due to bedrock lithology could be inferred. Watersheds with more abundant supplies of sediment (e.g. East Fork San Juan) had relatively high rates of transport at bankfull flow (*c*. 0·025 kg s^{-1} km^{-2}) while other sites with lower supplies of sediment had more modest (0·002 kg s^{-1} km^{-2}) or low (0·0001 kg s^{-1} km^{-2}) rates of transport. There was a general and predictable increase in the rate of transport moving from small to large channels, as demonstrated using sites from the Fraser Experimental Forest. Some of the inherent variability in transport rates could be attributed, in part, to seasonal influences on the supply of sediment. However, there was no consistent pattern of hysteresis, indicating that seasonal variation in sediment flux is specific to conditions at individual sites.

Regardless of channel type or overall transport rate, the results from the piecewise regression analysis support a near-bankfull threshold for the onset of phase II transport. Other investigators have noted a rapid increase in the rate of sediment transport in coarse-grained channels, though not all refer specifically to transport phases (Milhous, 1973; Parker, 1979; Parker *et al*., 1982; Jackson and Beschta, 1982; Andrews, 1984; Carling, 1988; Ashworth and Ferguson, 1989; Warburton, 1992; Andrews and Nankervis, 1995; Whiting *et al*., 1999; Emmett, 1999). For instance, Carling (1988) concluded that onset of phase II transport begins at about 60 per cent of bankfull in an alluvial channel in steady-state equilibrium. Whiting *et al*. (1999) observed breakup of the armour layer and rapid increase in transport rates in Idaho streams beginning at about 80 per cent of bankfull, on average. Trush *et al*. (2000) suggest that significant gravel entrainment begins between 75 and 110 per cent of bankfull and that the range is related to within-channel differences in the grain size, morphology, and location of sediment deposits. Hence, there is substantial evidence in the literature supporting a threshold at which there is a change in the nature of bedload transport in many coarse-grained streams. One advantage of the piecewise regression analysis is that it presents an objective and statistically defensible means for determining this flow. Moreover, the results from this study support commonality in the onset of phase II transport across a range of channel types. It appears that a channel adjusted to the imposed sediment load and flow begins to substantially mobilize sediment comprising the channel surfaces at flows approaching bankfull. The actual rates and sizes transported, however, are a function of the flow and sediment supplied to the

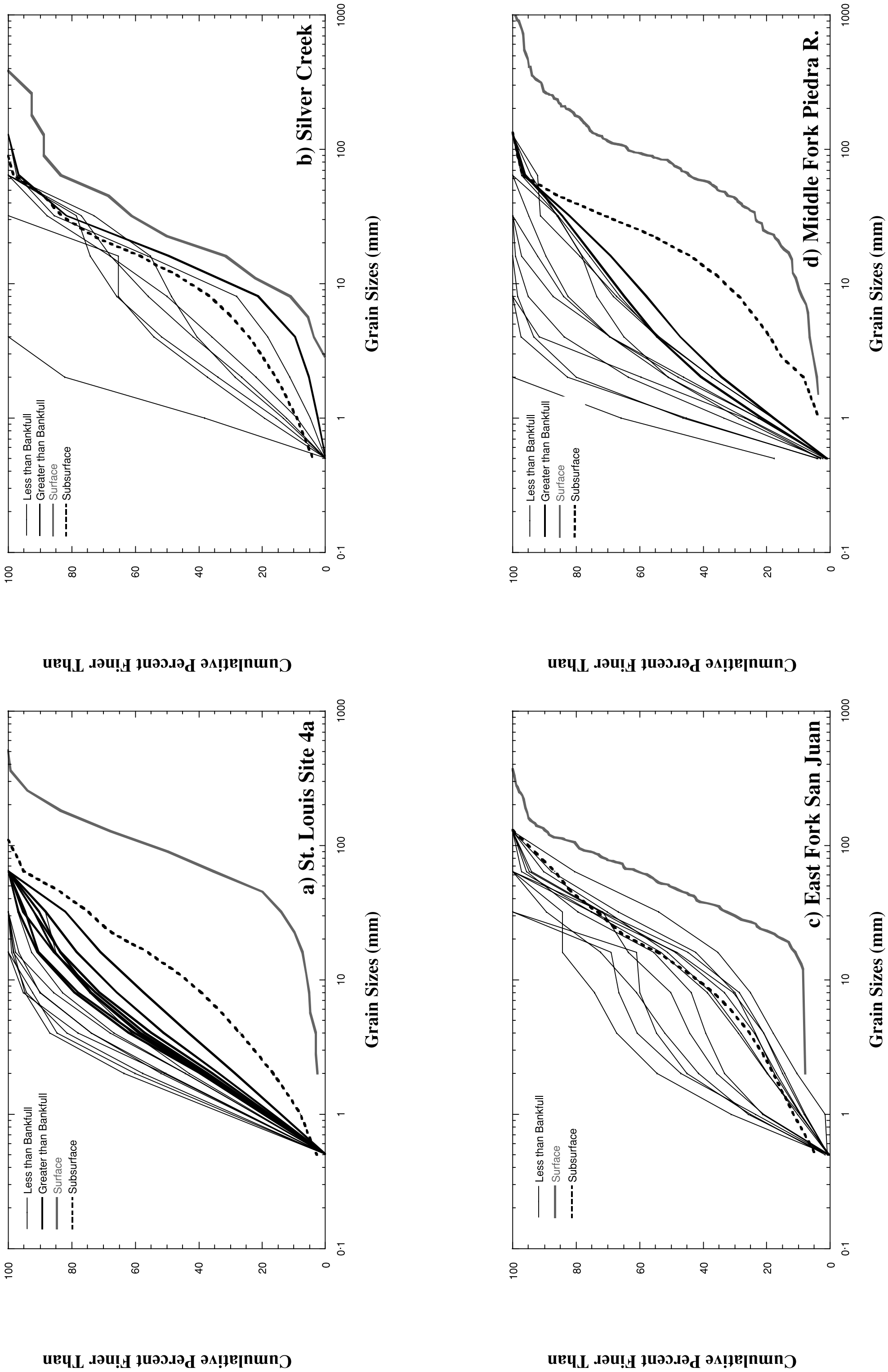

Figure 9. Comparisons of grain size distribution for bedload (solid black lines), subsurfaces (dashed line), and bed surfaces (solid grey line) for four sites. Bedload distributions are from composites of samples obtained over narrow flow ranges (e.g. 0–5 per cent bankfull, 5–10 per cent bankfull). Thicker black lines are from flows that are greater than bankfull while thinner black lines are from less-than-bankfull flows. The curvilinear patterns for other sites (not shown) are similar to that depicted in (a) and (d). The patterns shown for (b) Silver Creek and (c) East Fork San Juan indicate that bedload is coarser than the subsurface at relatively moderate discharges and becomes similar to that of the surface at flows approaching bankfull.

system. Substantial departures in the value of the relative breakpoint could be due to several factors, including that the system is in a state of disturbance.

While there were between-site differences in grain sizes collected beginning at the breakpoint discharge, there was consistency in the relative size of these grains when scaled by the bed surface material. The surface D_{16} to D_{25} appears to be an important grain size whose capture in bedload samplers represents the onset of phase II transport in most of our gravel-bed channels. However, this result may only hold for channels where the armour layer is largely static or stable over a range of flows, as it is at most of our sites. By contrast, grains representing about D_{60} of the channel surface were collected at relatively low discharges at two sites. Here, the grain size distribution of the bedload became coarser than that of the subsurface at less than bankfull discharge and approached that of the surface at near-bankfull flow. The movement of coarser grains over a wider range of flows suggests differential mobility of the channel surface at these two sites.

While the 19 sites described herein represent a range of gravel-bedded channels that lie on a continuum from stable step–pool systems to channels with more mobile beds and braided configurations, most exhibited two phases of bedload transport. These results confirm those of our earlier study (Ryan *et al*., 2002) that a change in phase of bedload transport occurs over a narrow range of discharge typically between 60 and 100 per cent of bankfull.

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